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The principal part of this report concerns the study of Lg phases in France. In a previous work, we obtain anelastic attenuation coefficient and we tried to model regional phases in order to understand the influence of the source. We show here that it is possible to derive directly from the records, not only the attenuation but simultaneously the source function and the transfer function at each station. We also present in this report, the first part of a signal processing, which will enable to improve the magnitude determination of quakes as well as the transfer function of our network for teleseisms.

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STATIONS CORRECTION AND COMMON SIGNAL

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A - STUDY OF REGIONAL PHASES

I - Propagation and excitation of regional phases in western Europe

Recall of former work.

In a first approach (1980), we have evaluated the attenuation due to geometrical spreading and anelasticity of recorded regional phases for a set of selected earthquakes which occurred in France or around the French territory. (Nicolas and al. BSSA vol. 72 n°6 pp. 2089-2106 December 1982) This purely experimental study led to the conclusion that a global attenuation factor in the range of distances from 100 to 1000 km could be defined on broadband data (0.5 - 16 Hz). Global attenuation coefficients γ ($A \propto D^{-\gamma}$) around 2 for P_n and S_n , and around 2.3 or 2.5 for P_g and L_g were obtained. Assuming an attenuation due to geometrical spreading α around 0.5 ($A \propto (D \sin D)^{-\alpha}$) we could compute the anelastic attenuation coefficient K ($\propto e^{-KD}$). For L_g waves propagating in western Europe, K is of the order of $0.2 \pm 0.1 \text{ deg}^{-1}$ to compare to 0.07 deg^{-1} for Eastern US and 0.5 deg^{-1} for northern USSR, or 0.35 deg^{-1} for the south of Caspian sea (Nuttli).

It was also checked that the quality factor Q was depending on frequency for each of these phases P_n , S_n , P_g , L_g . Even if an accurate computation was not realistic, it was possible to estimate Q around some 100 to some 1000 when the frequency increases from 1 Hz to 16 Hz.

In a second approach (1982), we tried to model regional phases by using the Bouchon's method, in order to understand the influence of the source (depth, source mechanism) and of the propagation model on the seismograms built at various distances. The Bouchon's method has the advantage of a full waves method, which computes exact seismograms (Campillo and al BSSA vol. 74 n°1 pp. 79-90, February 1984).

In addition this theoretical study brings an estimation of the geometrical attenuation with epicentral distance more accurate than the previous one: for L_g waves of the form $D^{-0.83}$; for P_g waves of the form $D^{-1.5}$. The excitation of P_g wave is found insensitive to the depth of the source within the crust while the L_g wave amplitude is found 50 per cent higher for a source in the upper and middle crust than in the lower crust. The

amplitudes of these two phases drastically decrease when the source is below the Moho, which underlines the important role of wave guide played by the crust for the propagation of P_g and L_g . We also find that the geometrical attenuation of P_g and L_g waves is independent of source mechanisms. In the case of an explosion, the excitation of P_g is insensitive to the source depth. The L_g wave amplitude is small in comparison to P_g and L_g waves and depends on the closeness of the source to an interface or to the free surface.

II - Research and development, during 1984 by data analysis of L_g phases

Taking into account the last results obtained on the attenuation of L_g waves due to geometrical spreading, we have made an attempt to improve our former evaluation of the anelastic attenuation and consequently of the Q factor. The method we have used also leads to the evaluation of the source spectrum of the quake and of the station response.

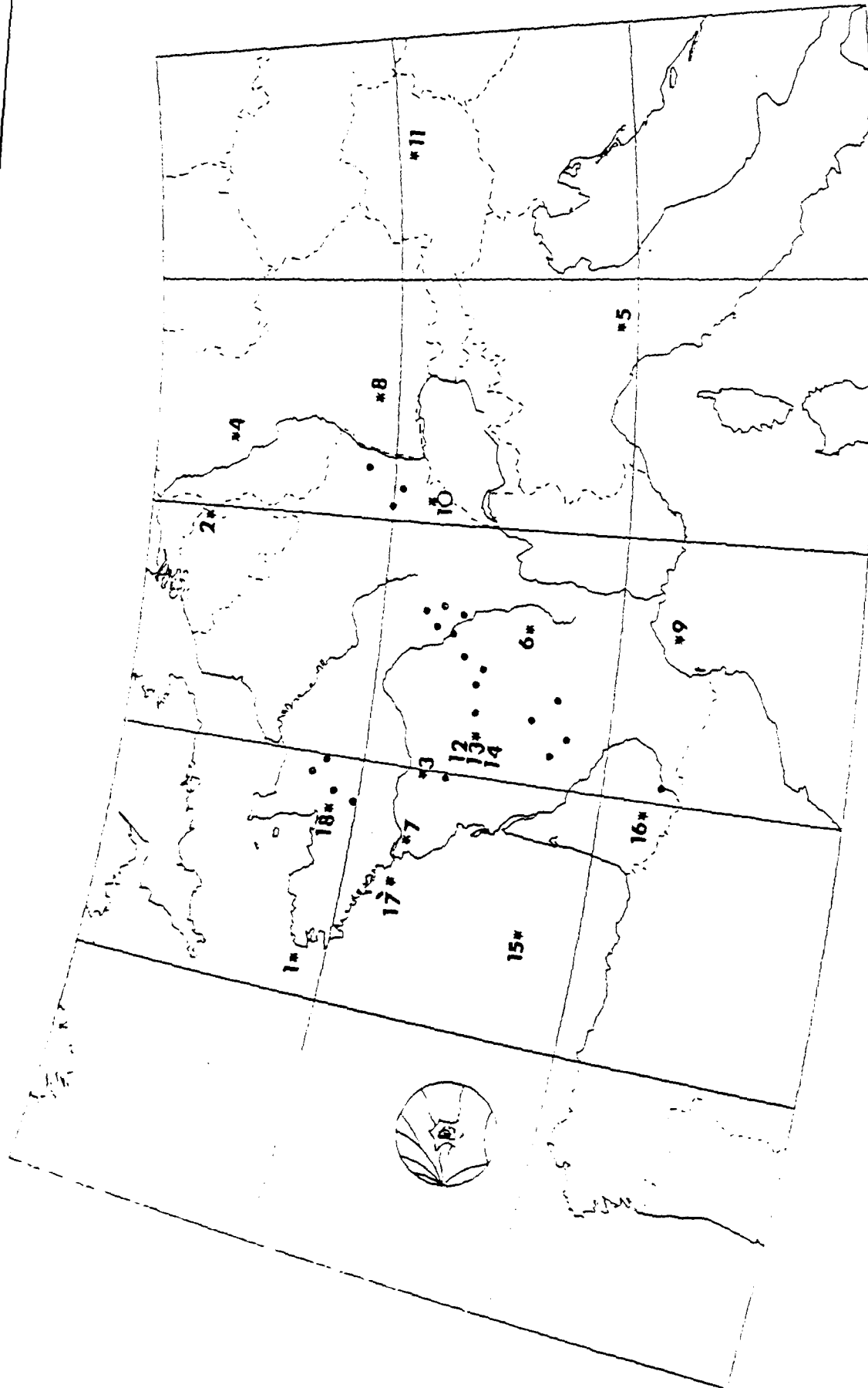
The L_g phases which are made of multireflected S waves within the crust (Bouchon 1982) carried on informations on the source, the propagation and the reception. In the frequency domain, the L_g amplitudes are expressed as the product of these three terms : Source, Attenuation and Reception.

We have separated the influence of these three terms by studying L_g spectra recorded on the french seismic network of LDG.

This network of large aperture (800km x 800km) is well adapted to that study. The telemetered data from its 28 SP seismographs (vertical component) are sampled at 50s. per sec. and recorded on digital form. Consequently, spectra study from 0.5Hz to 15Hz is possible with a reasonable definition.

1°) Data and Data processing

L_g waves records of 18 earthquakes located around France (Fig.1) with magnitudes ranging from $M_L = 3.2$ to $M_L = 4.8$ are selected (Table I). Propagation paths range from 150km to 1200km but for one quake the maximum recording distance is generally two or three times the minimum distance due to the dynamic range (60db) of the digital recording. Each L_g waves train is divided in three parts according to their group velocities (Fig.2).



° STATION
 15* EARTHQUAKE (The number refers to Table I)
 Figure 1

	DATE	ORIGIN TIME	COORDINATES	ML	REGION
1 -	04/00/81	04h 41 59.2	48.48N 05.11W	4.2	BREST
2 -	02/03/82	01h 27 26.3	51.01N 05.77E	4.0	ANVERS
3 *	28/05/82	04h 50 24.7	46.98N 00.11W	3.3	LA ROCHELLE
4 -	28/06/82	09h 57 33.3	50.69N 07.84E	4.9	FRANKFURT
5 -	26/07/82	15h 07 29.5	44.26N 10.88E	4.3	EMILIE
6 *	08/10/82	13h 19 46.3	45.51N 03.66E	3.3	CLERMONT-FERRAND
7 *	09/11/82	13h 44 47.3	47.07N 01.73W	3.9	NANTES
8 -	28/11/82	04h 34 05.0	48.31N 09.05E	3.9	JURA SOUABE
9 *	23/12/82	14h 48 13.4	43.02N 03.75E	4.1	MONTPELLIER
10 -	03/02/83	02h 48 30.1	47.34N 06.53E	3.4	VESOUL
11 -	14/04/83	14h 52 15.2	47.75N 15.10E	4.8	SALZBURG
12 *	21/04/83	01h 53 07.9	46.20N 00.96E	4.0	BELLAC
13 *	21/04/83	19h 07 02.1	46.19N 00.98E	3.6	BELLAC
14 *	21/04/83	23h 31 13.8	46.20N 00.98E	3.8	BELLAC
15 -	08/05/83	17h 47 51.4	44.97N 03.47W	4.0	OUEST-ROCHEFORT
16 -	06/06/83	01h 29 50.3	43.27N 00.30W	4.1	PYRENEES
17 *	03/07/83	20h 47 11.2	47.21N 02.80W	3.2	LORIENT
18 -	07/07/83	03h 52 25.2	48.41N 01.30W	4.2	FOUGERES

- CSEM LOCATION

* LDG LOCATION

ML LOCAL MAGNITUDE (LDG)

TABLE I

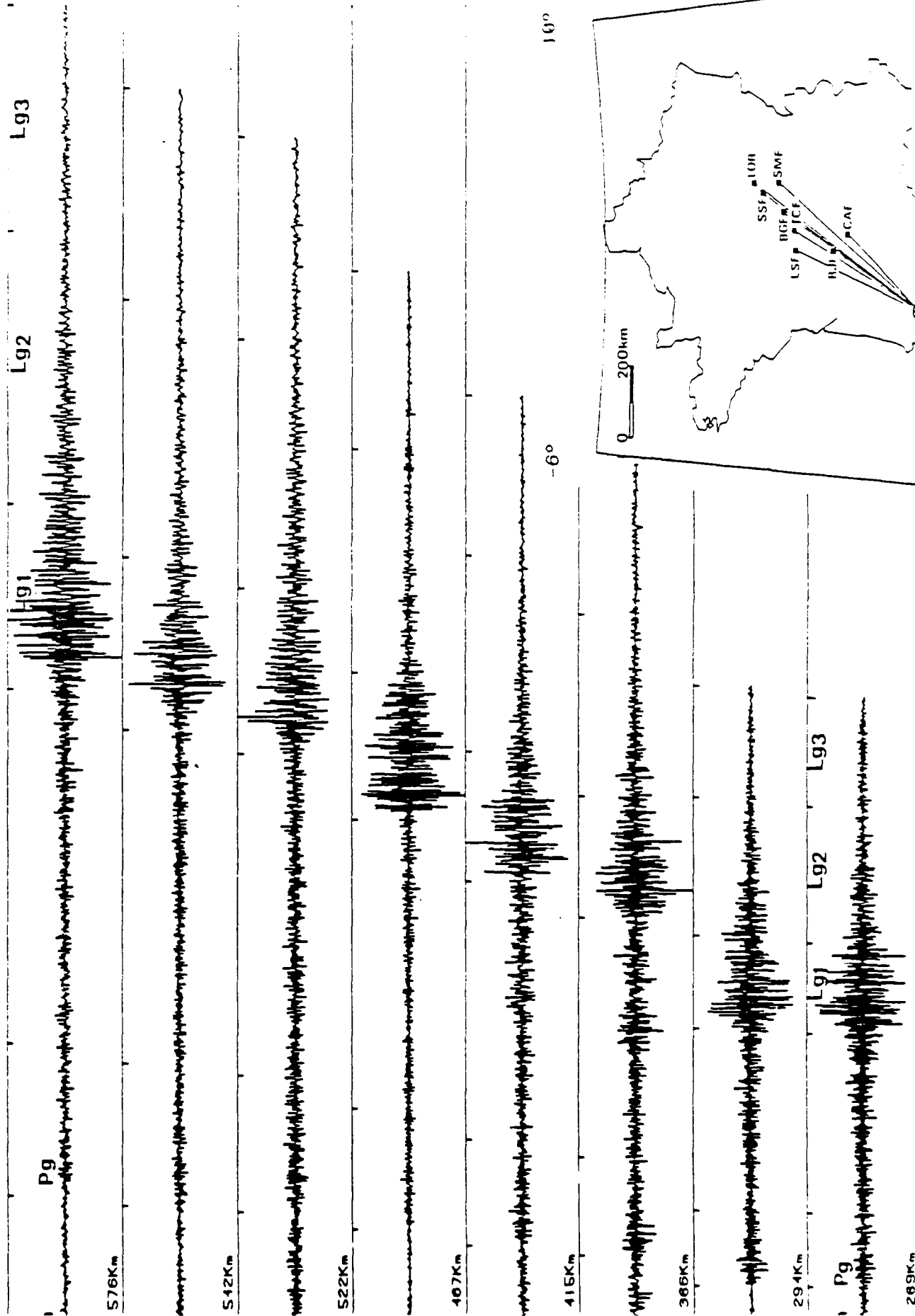


figure 2 : Seismograms recorded in different stations of the french network for a pyrenean quake (n°16).
See L_{g1} , L_{g2} , L_{g3}

$$L_{g1} = 3.6 \text{ km/s} > V > 3.1 \text{ km/s} \text{ maximum amplitude}$$

$$L_{g2} = 3.1 \text{ km/s} > V > 2.6 \text{ km/s} \text{ begining of coda}$$

$$L_{g3} = 2.6 \text{ km/s} > V > 2.3 \text{ km/s} \text{ end of coda}$$

For each L_g part, the frequency spectrum is computed (Fig.3) within 0.5 and 15Hz.

Each amplitude $A_i(f,d)$ for the i^{em} station and j^{em} quake is expressed as follows

$$A_{ij}(f,d) = S_j(f) * EXG(d) * AI(f,d) * FTS_i(f) \quad (1)$$

where

- $S_j(f)$ is the source spectrum of the quake j .

- $EXG(d)$ is the attenuation term for geometrical spreading.

- $AI(f,d)$ is the term for anelastic attenuation.

- $FTS_i(f)$ is the response function of the station i .

f is frequency d is distance

with
$$AI(f,d) = e^{-\frac{\pi f d}{Q V}}$$

V = group velocity

Q = quality factor

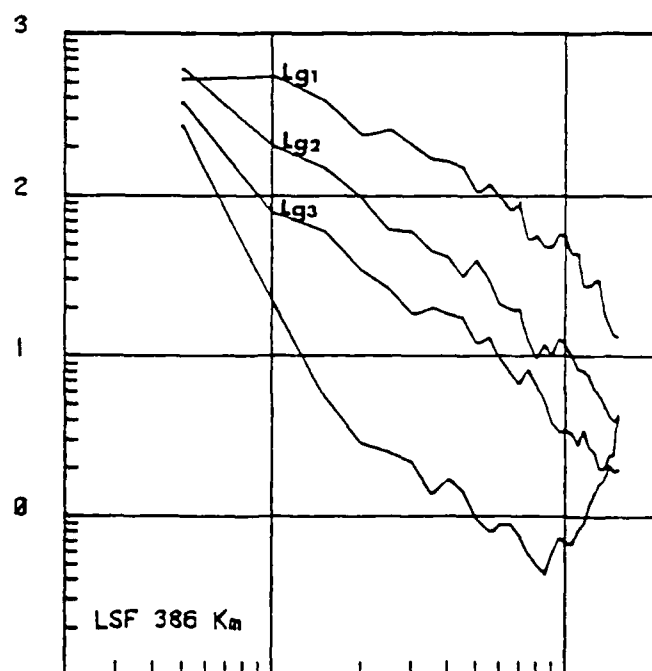
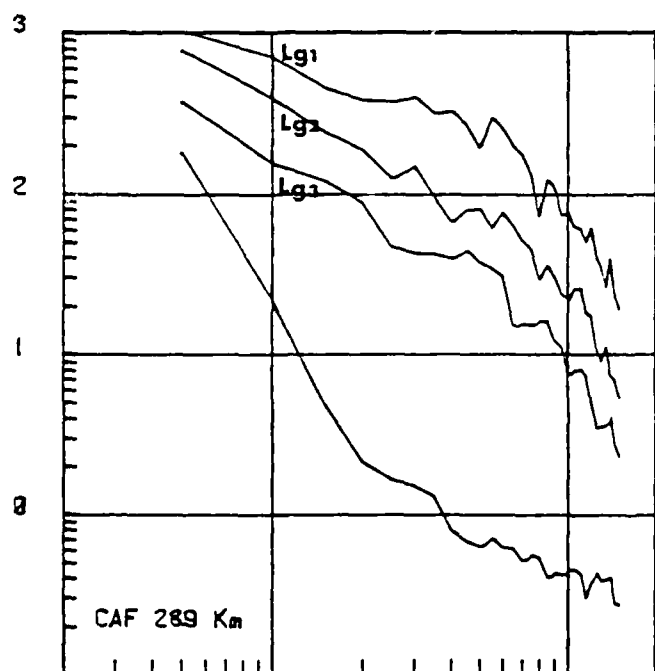
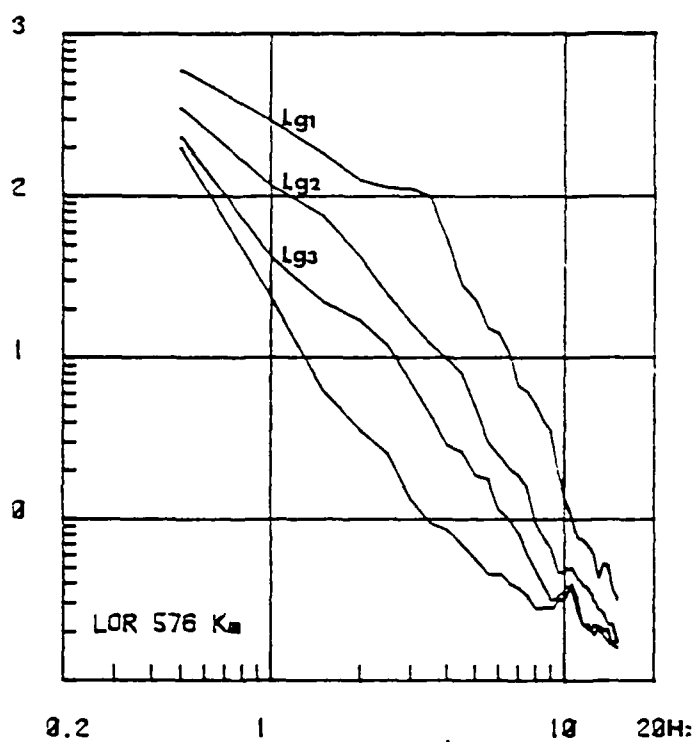
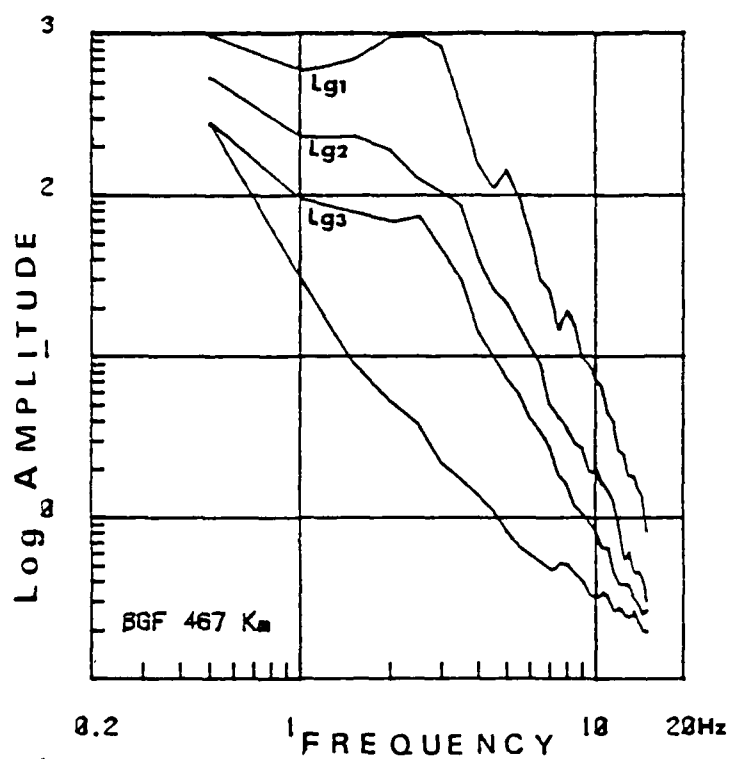
The geometrical attenuation term has been found previously of the form $EXG(d) = d^{-0.8}$

Consequently equation(1) is :

$$\log A_{ij}(f,d) - 0.8 \log d = \log S_j(f) - \frac{\pi f d}{Q V} + \log FTS_i(f) \quad (2)$$

which is solved by iteration process.

Figure 3 : Frequency spectra of L_g waves for different epicentral distances.
(Seismic noise corresponds to the lowest one)



a) For each frequency and each quake $S_j(f)$ and $Q(f)$ are estimated by least squares ($FTS_i(f)$ is put to 1).

b) For the whole set of earthquakes we build up a relation

$$Q = Q_0 f^b$$

c) For each station, we estimate the perturbation to bring to its transfer function as the mean value of the perturbations given by all the earthquakes in this station, with the additive hypothesis

$$\sum_{i=1}^n FTS_i(f) = 1$$

this process is stable and convergent within few iterations.

2°) Main results

a) Transfer function at the i^{em} station.

By taking into account the station response, the Q factor estimation for L_g is significantly improved and the data dispersion is reduced. The transfer functions for the 23 studied stations correspond essentially to attenuation or amplification by a factor up to 3 for high frequencies ($f > 5\text{Hz}$) Fig. 4 .

These transfer functions can be explained by important variations of Q factors associated with superficial layers beneath the station area. They are well correlated with variations of the seismic noise spectrum recorded in the station, precisely for high frequencies which seems to confirm their superficial origine.

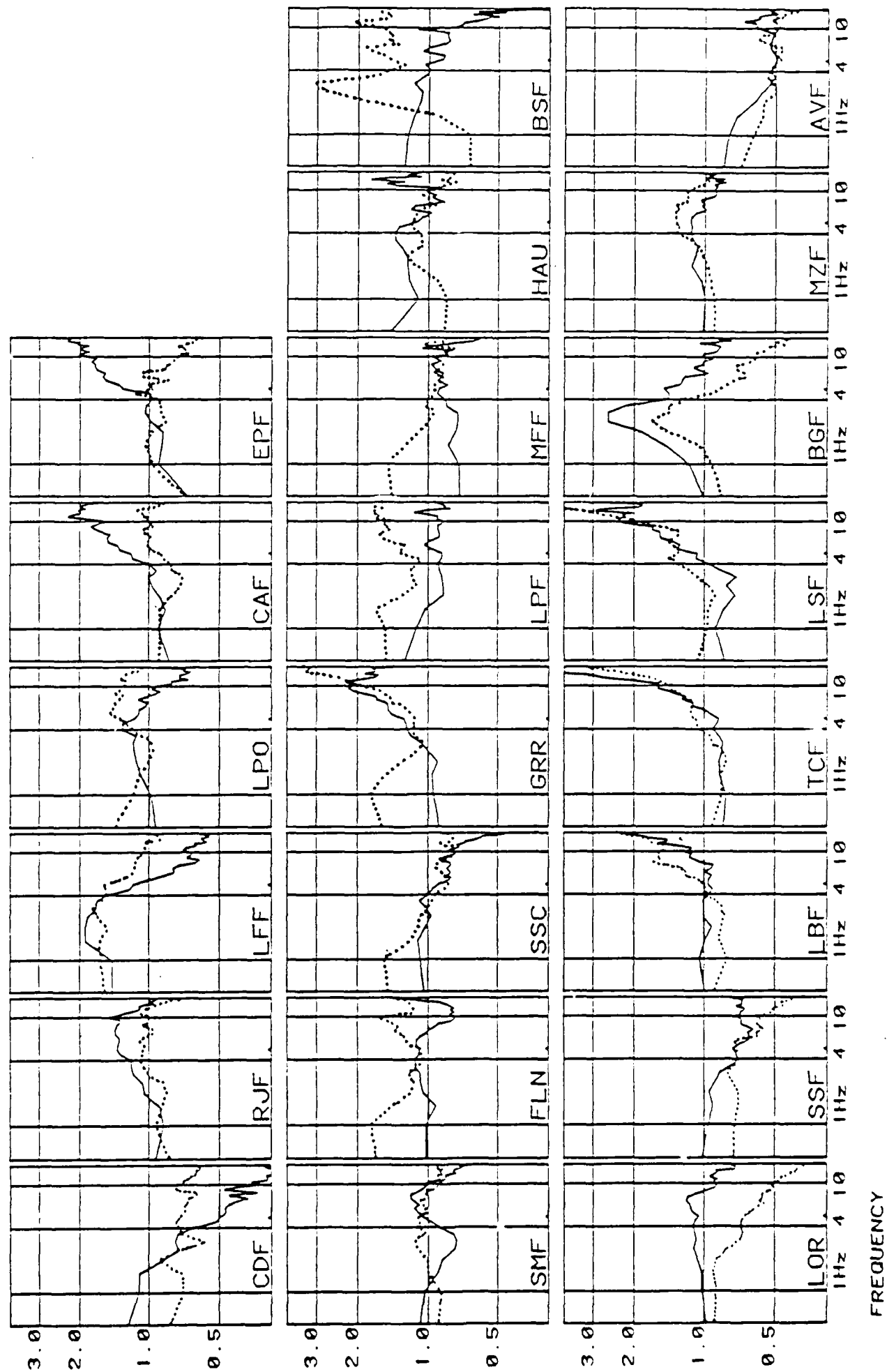
b) Quality factor

It has been computed independantly at each frequency for each earthquake. After removing the station transfer function, the data lead to an estimation of the Q factor with an accuracy of the order or better than 30% .

For all the quakes and the three L_g groups, the Q factor varies significantly with frequency following a relationship as

$$Q = Q_0 f^b \quad (\text{Fig. 5})$$

Figure 4 : Transfer function of each station (continue line). seismic noise spectrum (dotted line).



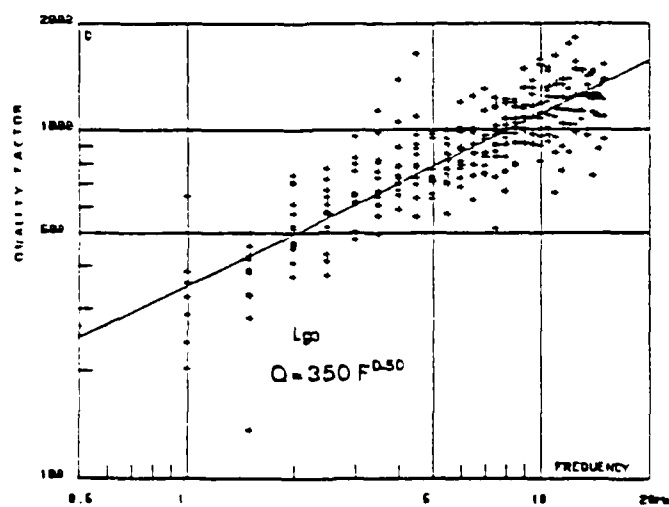
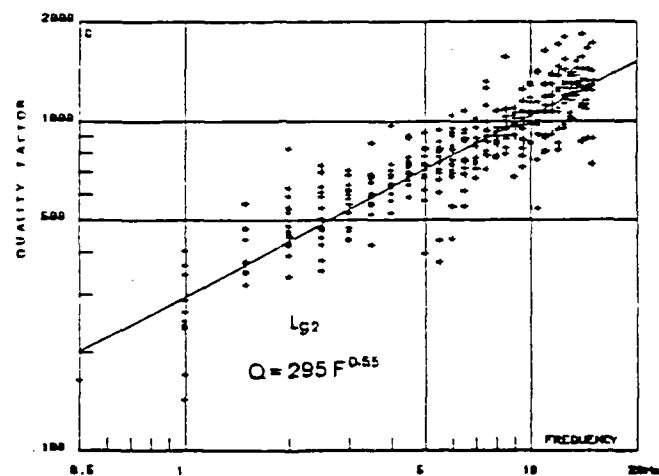
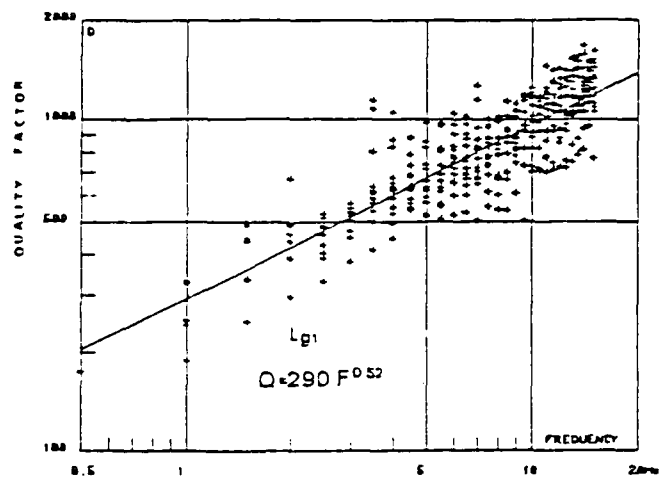


Figure 5 : Q factor versus frequency for L_{g1} , L_{g2} , L_{g3}

we obtain

$$\text{for } L_{g1} \quad Q = 290 f^{0.52}$$

$$\text{for } L_{g2} \quad Q = 295 f^{0.55}$$

$$\text{for } L_{g3} \quad Q = 350 f^{0.50}$$

These similar relationships for L_g phases and their coda agree with Aki hypothesis (1980).

c) Source spectrum

Each source spectrum is obtained as the mean of the spectra obtained in all the stations for the considered earthquake, corrected from each station transfer function, and from the propagation or attenuation term. The anelastic attenuation term being the one obtained globally as described in b). The spectra we have obtained have a constant dispersion probably due to a too global propagation term which does not take into account differences between the propagation structures.

Nevertheless they own the main general features of the Brune source. The L_g phases (L_{g1}) and their coda (L_{g2} and L_{g3}) lead to similar source spectra (Fig. 6). In the high frequencies domain of the spectra, the asymptotic decrease well defined for the most important quakes is between

$$\propto f^{-1.5} \quad \text{and} \quad \propto f^{-2.25}$$

The cut-off frequencies f_c between 1.25Hz and 10Hz are function of the magnitude M_L following the relationship

$$\log(f) = 1.98 - 0.33 M_L$$

On the other hand, the low frequencies level $S_0(0)$ of each source spectrum is a function of the magnitude M_L following:

$$\log S_0 = C + 1.5 M_L$$

true for L_{g1} , L_{g2} , and L_{g3} waves.

$$(\text{with } CL_{g2} = CL_{g1} - 0.38 ; \quad CL_{g3} = CL_{g1} - 0.68)$$

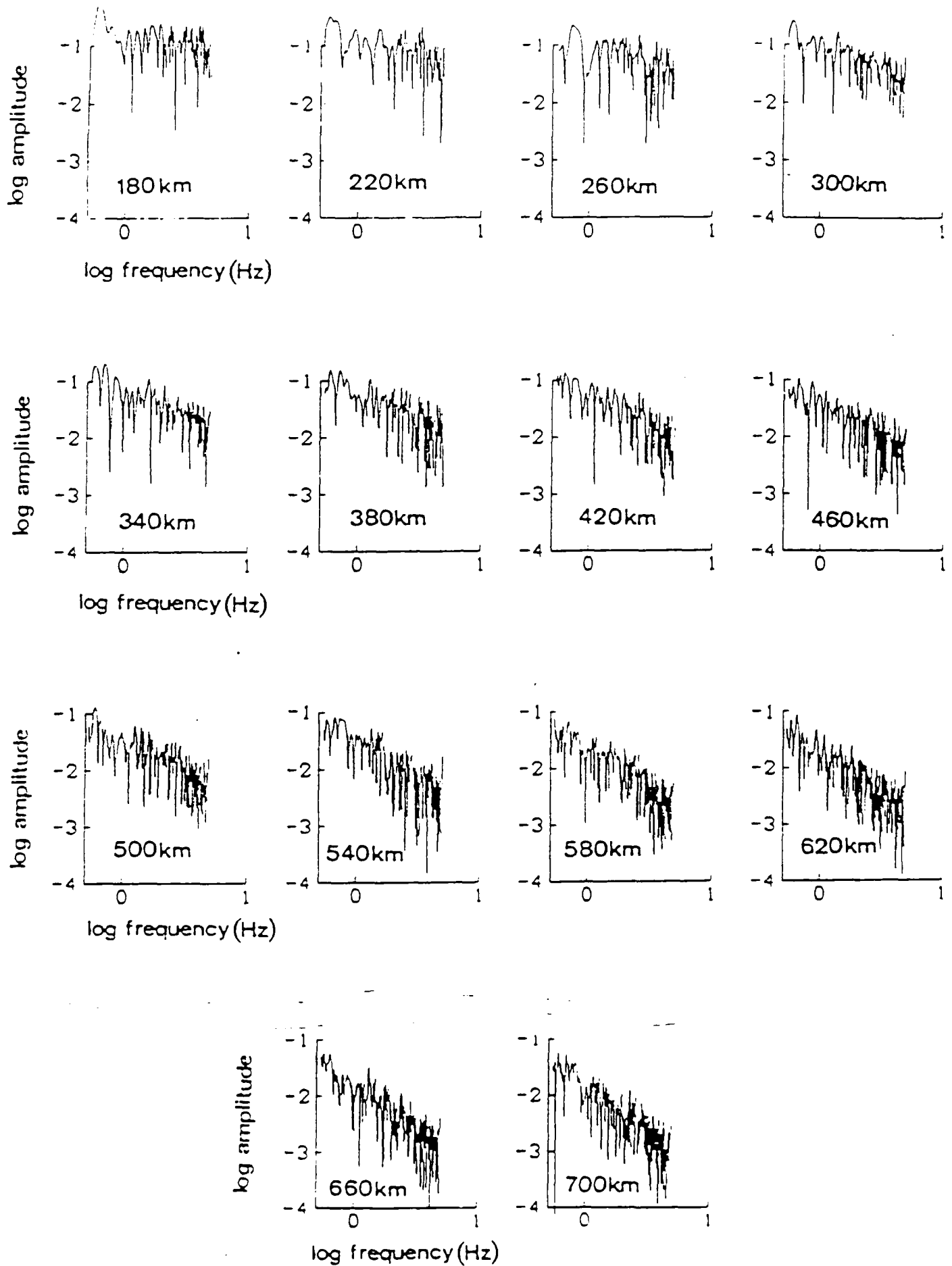


Figure 6 : L_g spectra obtained from synthetic seismograms

3°) Conclusion

Spectra of L_g phases and of their coda recorded in western Europe for 18 earthquakes have allowed us to evaluate the influence of the three essential effects which model them : source, anelastic attenuation, and station effects.

- The source spectra we have obtained have the main general characteristics of the Brune source.

- The quality factor Q similar for the L_g phases and their coda is depending on the frequency as we knew already :

$$Q \approx 300 f^{0.53}$$

It is a mean quality factor for western Europe.

- A transfer function for each station is estimated. It is well correlated to the seismic noise spectrum variations versus a mean spectrum over the network.

This transfer function mostly significant in the high frequencies domain is probably representative of the superficial layers attenuation under the station area.

III - Research and development in 1984: Numerical modeling of L_g phases

Taking into account the previous results both on geometrical spreading and anelastic attenuations of L_g waves in western Europe we have tried to interpret the Q factor of these waves as an effective measurement of Q_s , mean value of the quality factor of these waves in the crust. The crustal model that we use was inferred from long range refraction experiments in central France. It consists of a 4 layers crust overlying a mantle half-space.

For P waves we assume $Q_p = 2 Q_s$. The calculations are done using the discrete radial wave number representation of the seismic field (Bouchon 1981). The calculations are done for epicentral distances ranging from 180km to 700km and the results are shown in Fig.7 where they are compared to the case of a perfectly elastic crust. The source considered is a point of vertical strike-slip dislocation at a depth of 10km.

The seismograms represent the vertical component of the crustal response to a sudden step function dislocation. The frequency range extends

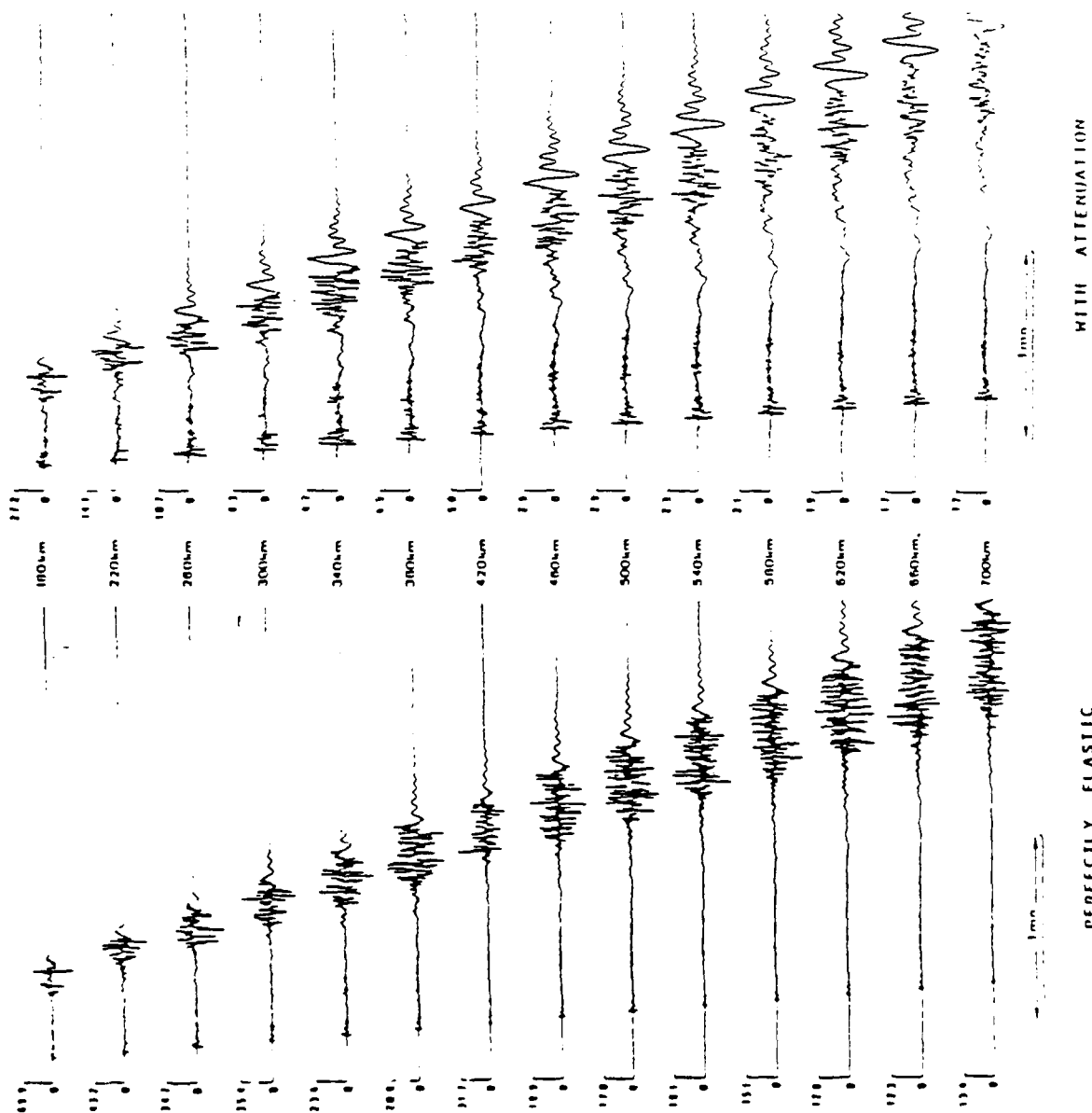


Figure 7 : Influence of attenuation on I_g phases (versus distance).

from 0 to 5Hz. The attenuation strongly affects the wave forms and the amplitudes and results in an enhancement of P_g waves and Rayleigh waves relatively to the L_g wavetrain.

The corresponding L_g wave spectra are presented in Fig.8 for a time window corresponding to group velocities between 3.8 km/s and 2.5 km/s. Attenuation of the high frequencies with increasing distances is clearly visible. At 180km, the effect of attenuation is small and the L_g wave spectrum is almost flat. This differs from a result we reported earlier (Campillo and al. 1984) and which was due to an incomplete evaluation of the wave-number series for large horizontal wave numbers at high frequency.

The numerical results obtained are compared with the data in Fig.9, after having normalized the theoretical spectra so that the mean values of observed and synthetic spectra at 180km are equal. The comparison with the data shows a very good agreement which supports the interpretation of L_g waves attenuation as a shear-wave crustal attenuation.

We have also looked at the effect of source depth and at the distribution of Q within the crust on the attenuation of L_g waves by ray tracing.

All the subcritically reflected rays have been drawn (Fig.10) radiated by sources located at 0 km, 10 km and 29 km depth.

Clearly the L_g waves radiated by the three sources sample with different weight different regions in the crust. A deeper source will sample the crust more uniformly than a source located in the upper crust.

We also have traced all the subcritically reflected rays radiated by a 10km deep source and leaving the source in three different ranges of take-off angles : $0 - 60^\circ$; $60^\circ - 120^\circ$ and $120^\circ - 180^\circ$. The near horizontal rays do not sample the lower crust while the more oblique one which are associated with longer travel time sample the crust quite uniformly (Fig.11).

Consequently, we have studied the sensitivity of L_g waves to a zone of strong attenuation in different regions of the crust. A zone of low Q in the sedimentary layers or at shallow depths has little effect on the L_g wave train with respect to the results of the perfectly elastic case (Fig.12a). A zone of low Q at the base of the crust on the contrary gives a sharp attenuation of the L_g wave coda (Fig.12b) which was insensitive to the presence of highly attenuation sediments.

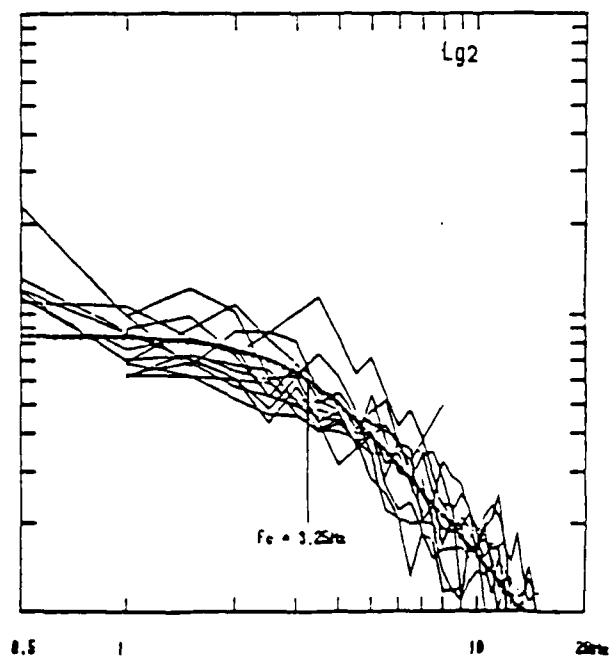
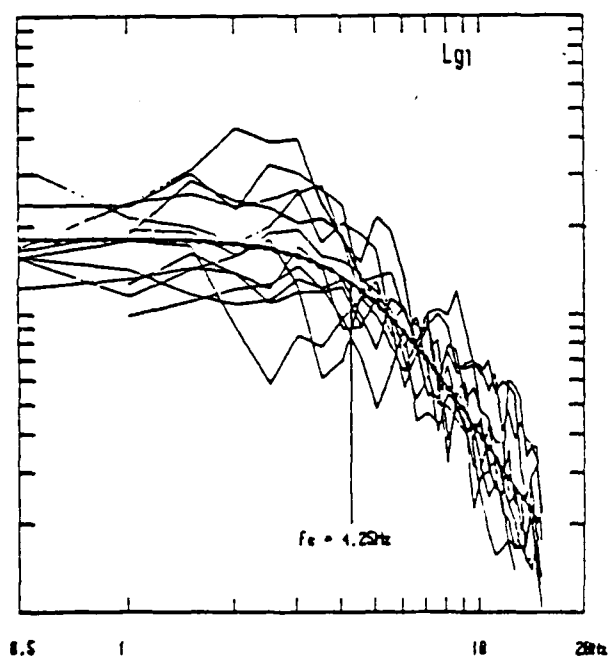
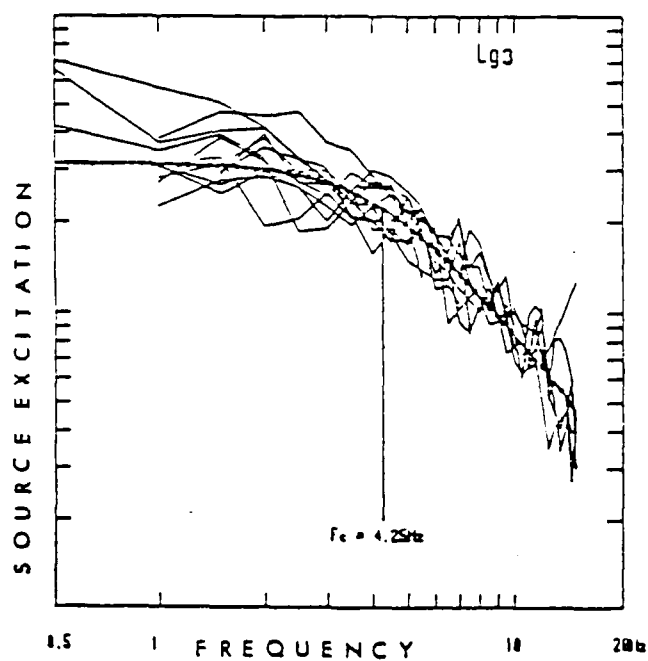


Figure 8 : Mean source spectrum of one quake.

Figure 9 : Energies of L_g waves versus epicentral distance at various frequencies :

- theoretical (continue line)
- data (dots)

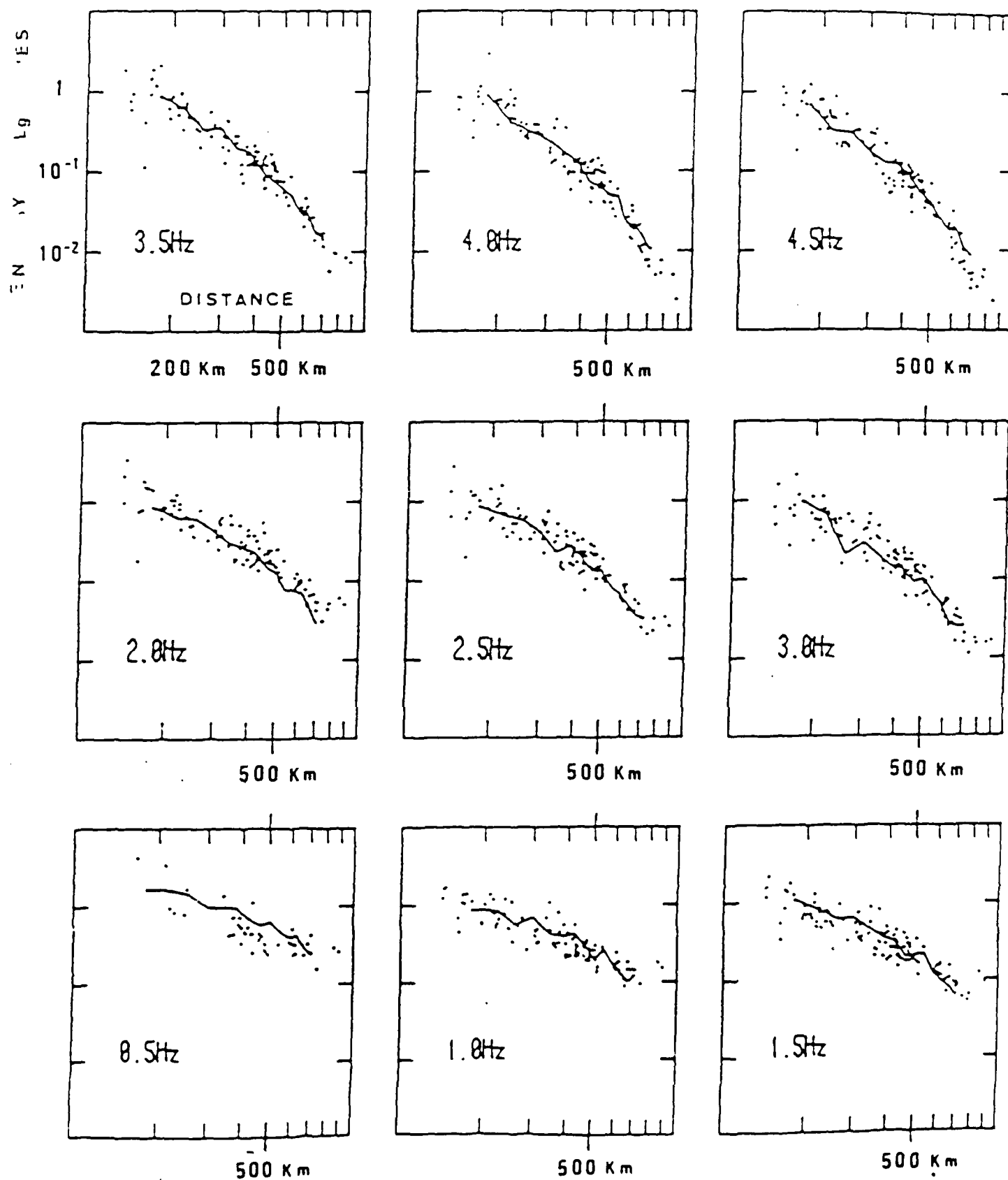


Figure 10 : L_g ray tracing for different source depths.

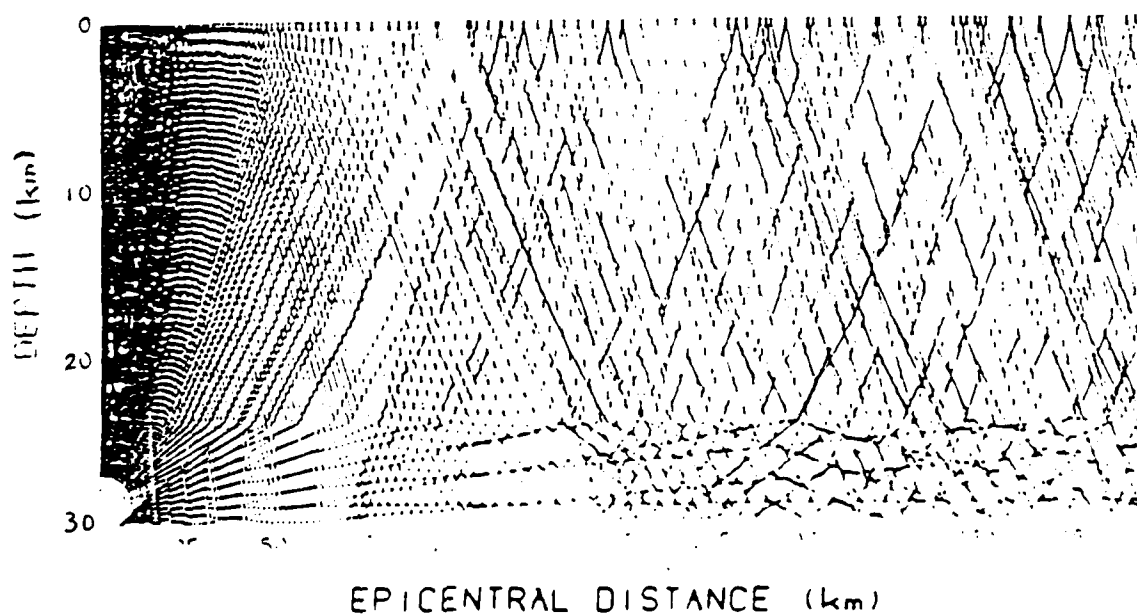
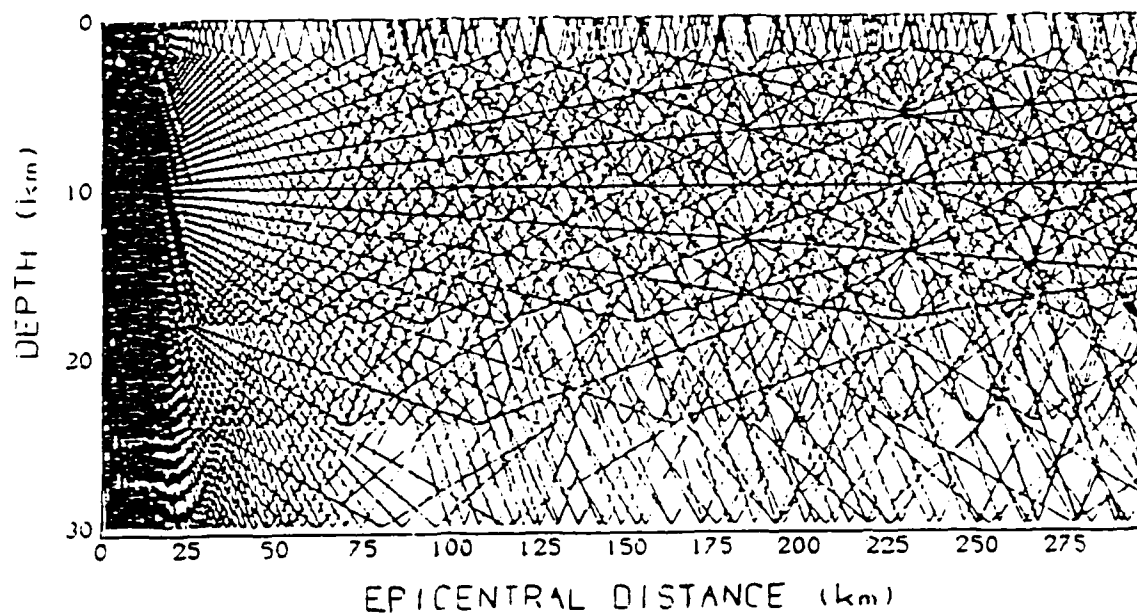
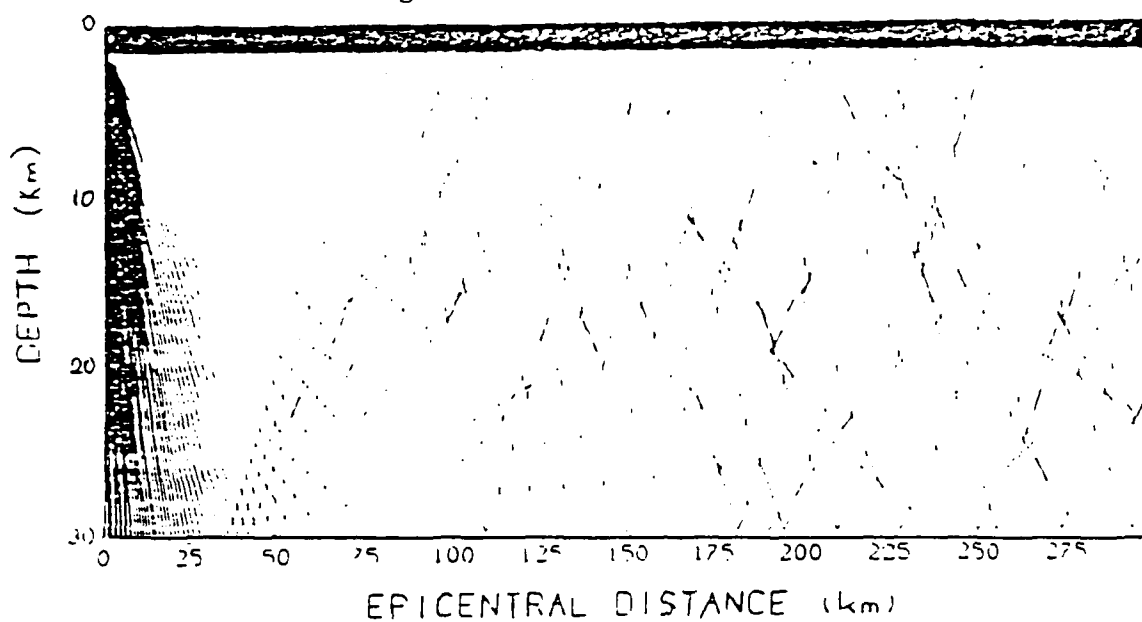
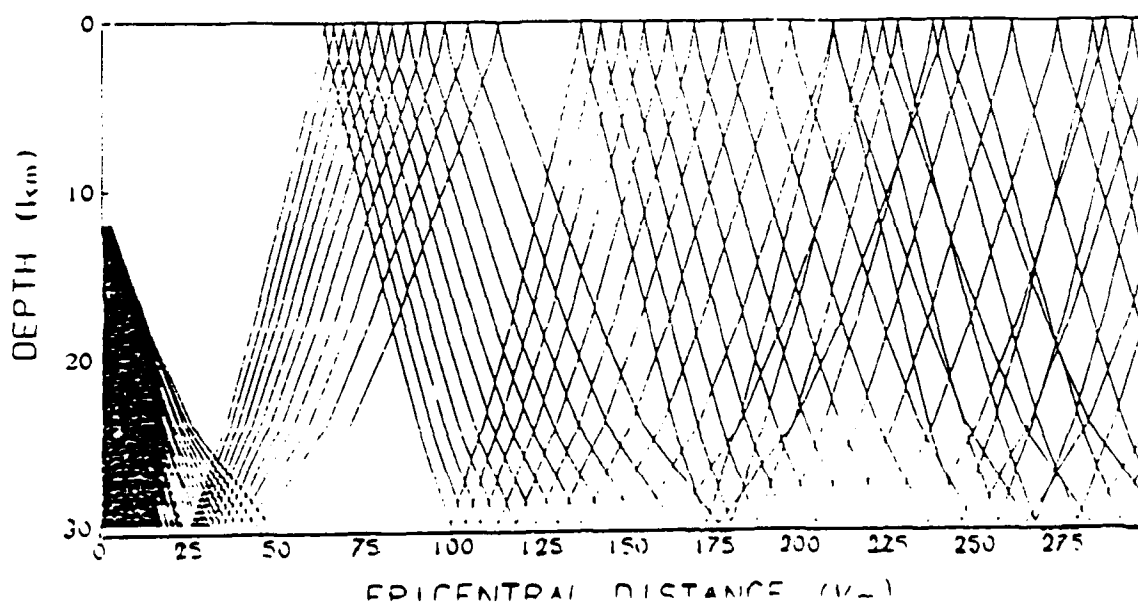
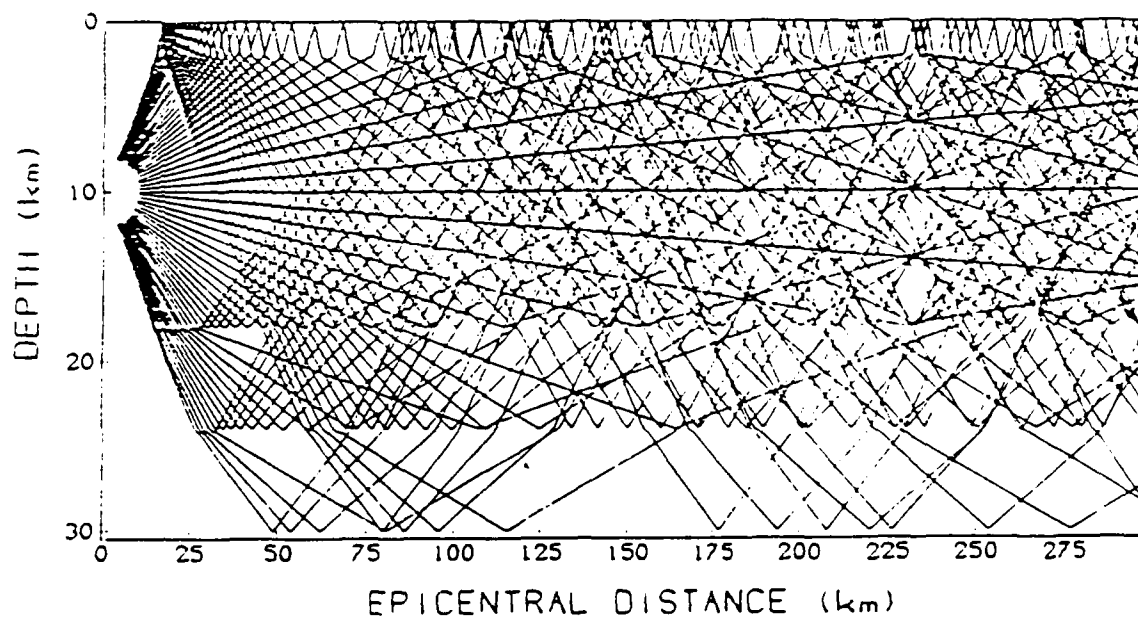
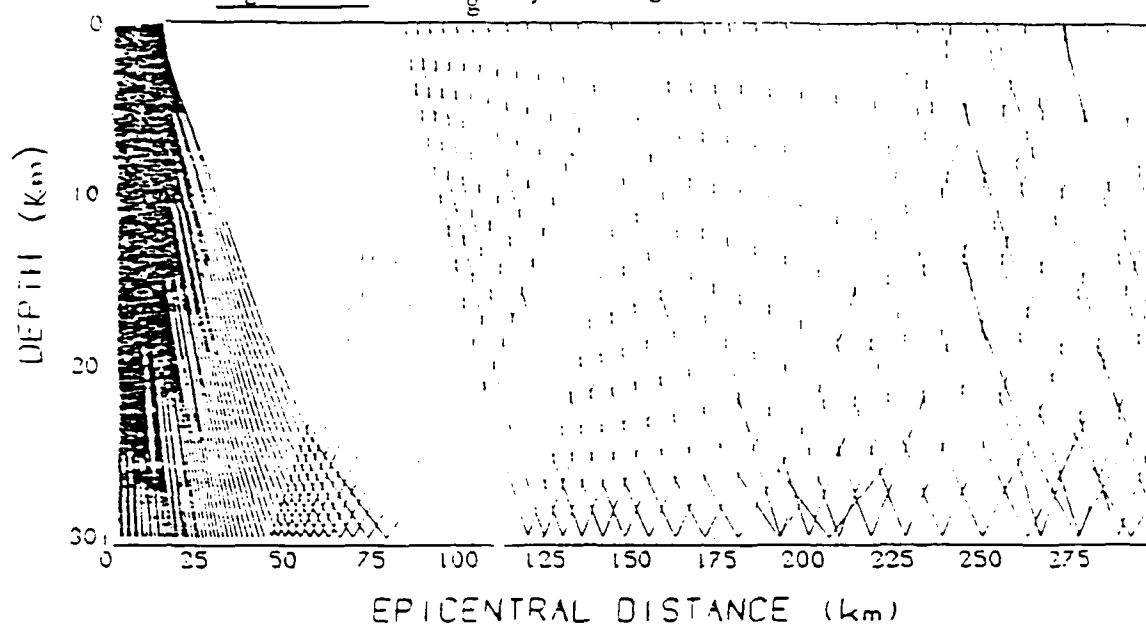


Figure 11 : L_g ray tracing for different ranges of take off angles.



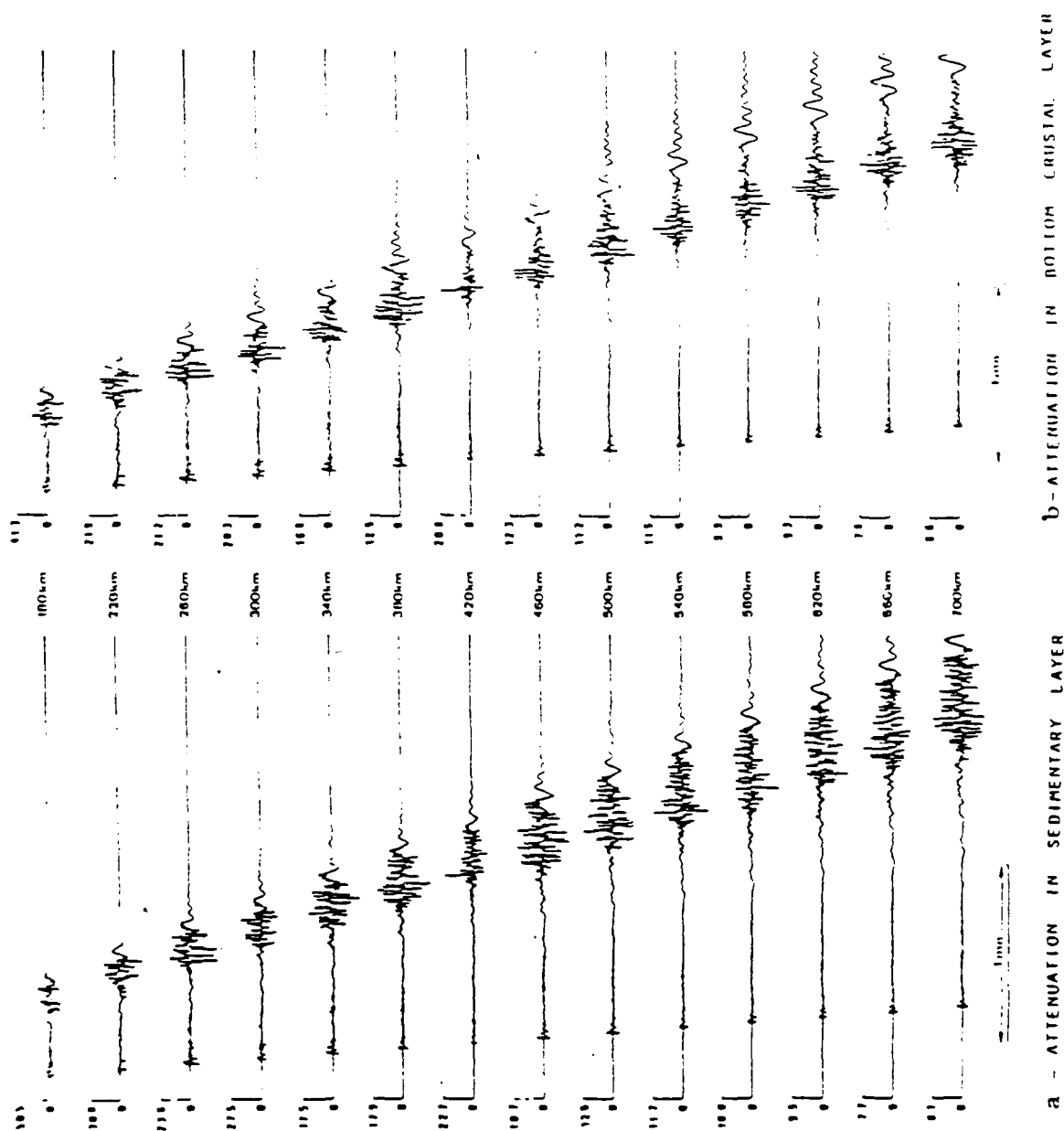


Figure 12 : Synthetic seismograms at various distances from the earthquake.

Coming back to the three different time windows L_{g1} , L_{g2} , L_{g3} for L_g waves trains recorded in France, the weak increase of the Q factor with decreasing group velocity indicates that Q factor of the lower crust is greater than the mean value of Q in the crust. Our results are compatible with the interpretation of the frequency dependency of Q as a tectonic indicator : stable regions are associated with Q which are monotonic and weakly increasing with frequency while, as pointed out by Aki (1980) active tectonic regions show a minimum of Q for frequencies around 0.5-1Hz which implies a stronger increase of Q with frequency as predicted by the theoretical works of Sato (1984).

IV - Future work

Excitation and propagation of regional phases and particularly of L_g phases have been studied theoretically up to now within homogeneous crustal layers.

Similarities between synthetics and recorded data came out although complexity of most of the phases is important.

We have seen that by using a more complicated propagation model (several layers, one sedimentary layer) we could approach in some cases a more realistic representation of the different phases.

We would like now to model the propagation of regional phases through heterogeneity zones such as large fracture zones, non planar layers etc....

This should be done by an extension of the discrete waves numbers method, with an analysis of the interface effect (Huygens Principe).

B - STATION CORRECTIONS AND COMMON SIGNAL

When an array receives a seismic signal, produced by a quake somewhere beneath the array, it is recorded in each of the stations, after a filtering process different for each of them and "polluted" by signal generated noise.

The aim of the processing we will describe is to invert this set of signals. Under ad-hoc hypothesis, it is possible to extract from the records, the common signal, the station corrections and the noise (mainly the coda of each record).

Data used in the analysis.

As a data processing is always very dependant of the kind of data used, let us first shortly describe the records which will be processed in the example given below.

The signals were recorded by the French Seismic Network ; 28 stations, short period vertical seismometers (1Hz) distributed over France : "Clusters" of stations (30 to 40km apart) the clusters themselves are at a distance, one to the other, of the order of 200 to 300km.

The events used in this analysis have to be relatively far away from France in order to record the same seismic phase in all stations. In the examples below, the epicenters are at a distance of the order of 40° from the centre of the network, in Central Asia.

We believe, the distance between the station to be relatively important in this analysis : not too small, nor too large.

Mathematical Analysis

At first, by taking into account the time of arrival, all records are phased. The station corrections, we will introduce, deal only with the amplitude of the signal and not with a possible time-term.

We will suppose those station corrections to be only a constant for each station (k_i for the i^{th} station). This could be considered as a crude approximation of the filtering of waves by the geological environment of the station. But it is not as drastic as it may seem : it is possible to work on narrow band filtered signals and to apply the same study for each frequency band.

Let us call :

- $e_{i,n}$, the n^{th} sample of the signal recorded in the i^{th} station.
- s_n , the n^{th} sample of the common signal we are looking for.
- $b_{i,n}$, the n^{th} sample of the noise in the i^{th} station (mainly, wave generated noise and coda).

We have to solve (in k, s and b) the equations

$$e_{i,n} = k_i \cdot s_n + b_{i,n}$$

In fact, s_n is not the common signal which we suppose the existence somewhere beneath the array, but this signal filtered by the mean geological structure of the array.

Those equations are strongly under-determined. The main constrain, we will add to solve them, is that everything in common between the records has to be put in the common signal. It means, using a least square approach that

$$X = \sum_i \sum_n (e_{i,n} - k_i \cdot s_n)^2$$

has to be minimized.

The problem may now be solve with only one under-determination. left : it is possible to multiply every k by the same factor if we divide the common signal s by the same constant. This fact will be expressed in an other way just below.

The normal equations

$$\frac{\partial X}{\partial k_i} = 0$$

$$\frac{\partial X}{\partial s_n} = 0$$

give

$$\sum_n e_{i,n} s_n = k_i \sum_n s_n^2 \quad \forall i$$

$$s_n = \frac{\sum_i k_i e_{i,n}}{\sum_i k_i^2} \quad \forall n$$

the last of them express that the common signal is a weighted sum of the records and allows us to eliminate s_n in the first set.

$$\left(\sum_j k_j^2 \right) \sum_n \sum_j k_j e_{j,n} e_{i,n} = k_i \sum_n \left(\sum_j k_j e_{j,n} \right) \left(\sum_\ell k_\ell e_{\ell,n} \right)$$

Introducing the cross correlation $C_{i,j}$ of the signals recorded in the i^{th} and j^{th} stations

$$C_{i,j} = \sum_n e_{i,n} e_{j,n}$$

we get

$$\left(\sum_j k_j^2 \right) \left(\sum_j C_{i,j} k_j \right) = k_i \sum_j \sum_\ell k_j k_\ell C_{j,\ell}$$

or, using the "bracket" notation :

$$|C|k\rangle = \frac{\langle k|C|k\rangle}{\langle k|k\rangle} \cdot |K\rangle$$

Obviously $|k\rangle$ is one of the eigenvectors of the matrix C . Its module is not defined, expressing in an other way the last under-determination we pointed above.

The expression X , to be minimized, can be written as

$$X = \sum_i C_{i,i}^2 - \frac{\langle k|C|k\rangle}{\langle k|k\rangle}$$

the eigenvalue associated to the eigenvector $|k\rangle$ has to be the largest one.

This analysis is very similar to the principal component analysis in statistic. It could be possible to expend the all set of records on the basis of all signal-vectors associated to each eigenvectors of the matrix C . But we only see a possible physical explanation for the first one, the common signal.

Examples :

We do not wish, here, to present neither the study of station corrections in our array versus azimuth or some other parameter, nor the study of a seismic swarm.

We will just give as an example of the processing on two quakes in Asia.

1° : 1982 Dec. 5 50°N 80°E $m = 6.1$

2° : 1983 Oct. 6 49.9N 78.8E $m = 6.0$

Those two events were chosen, among others, for their large magnitude and to be rather close one from the other. The focal mechanisms of both events are very similar.

The best is to first show some plotted signals.

The figure captions are :

Fig. 1 : Bottom part : 5 records of the first quake.

The total duration of the signals represented is 20 seconds. The amplitudes are normalized.

Upper part : records in the same stations for the second quake.

Fig. 2 : Common signals for both events.

Fig. 3 : For comparison : sum of the phased records for each event. The common signal is a weighted sum of records. The signals plotted on this figure are sum without weighting factors.

Fig. 4 : Signals and Coda : for both events, original signals in two stations and the computed coda : signal minus common signal (designated by C at the beginning).

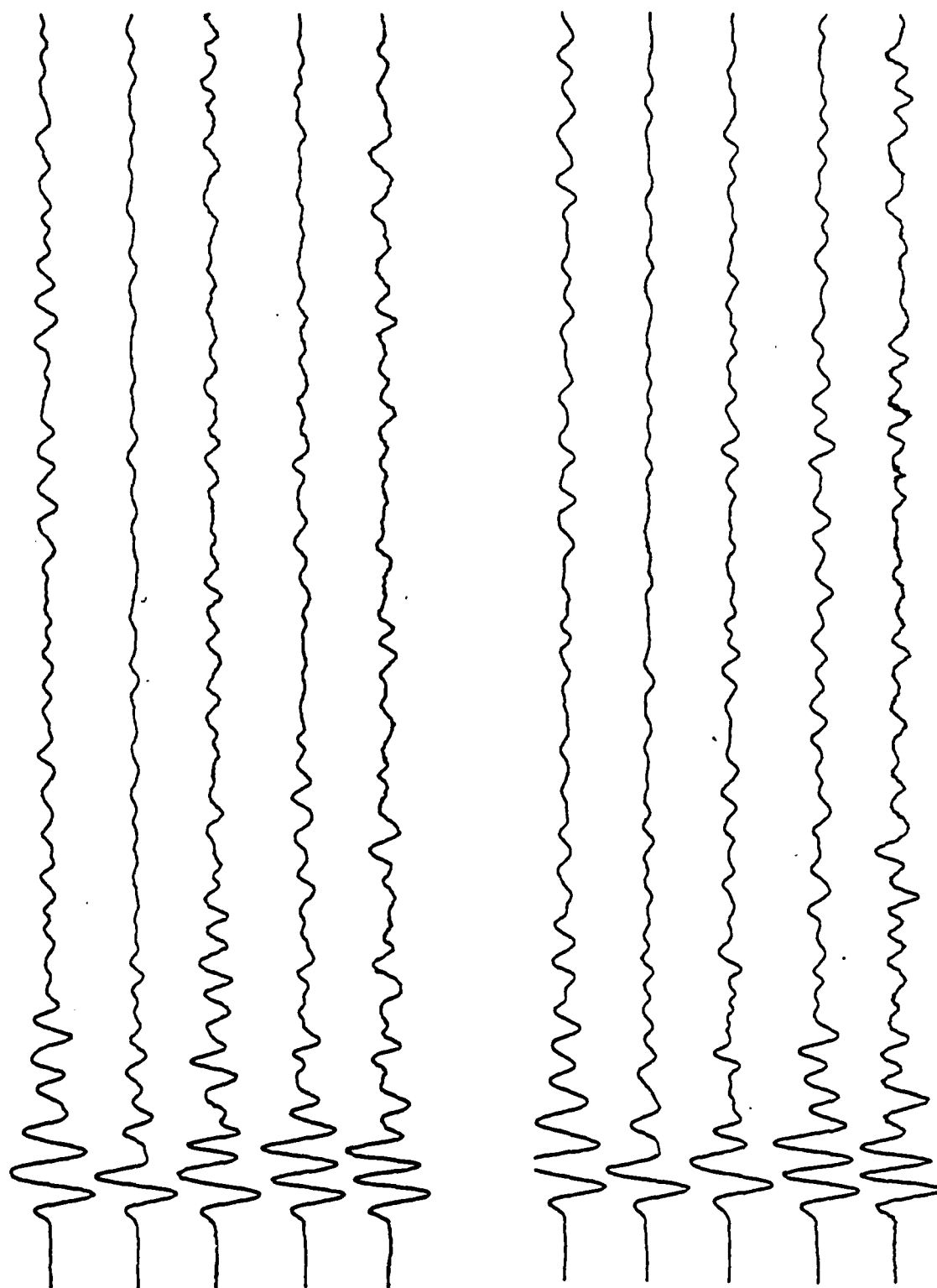


Figure 1

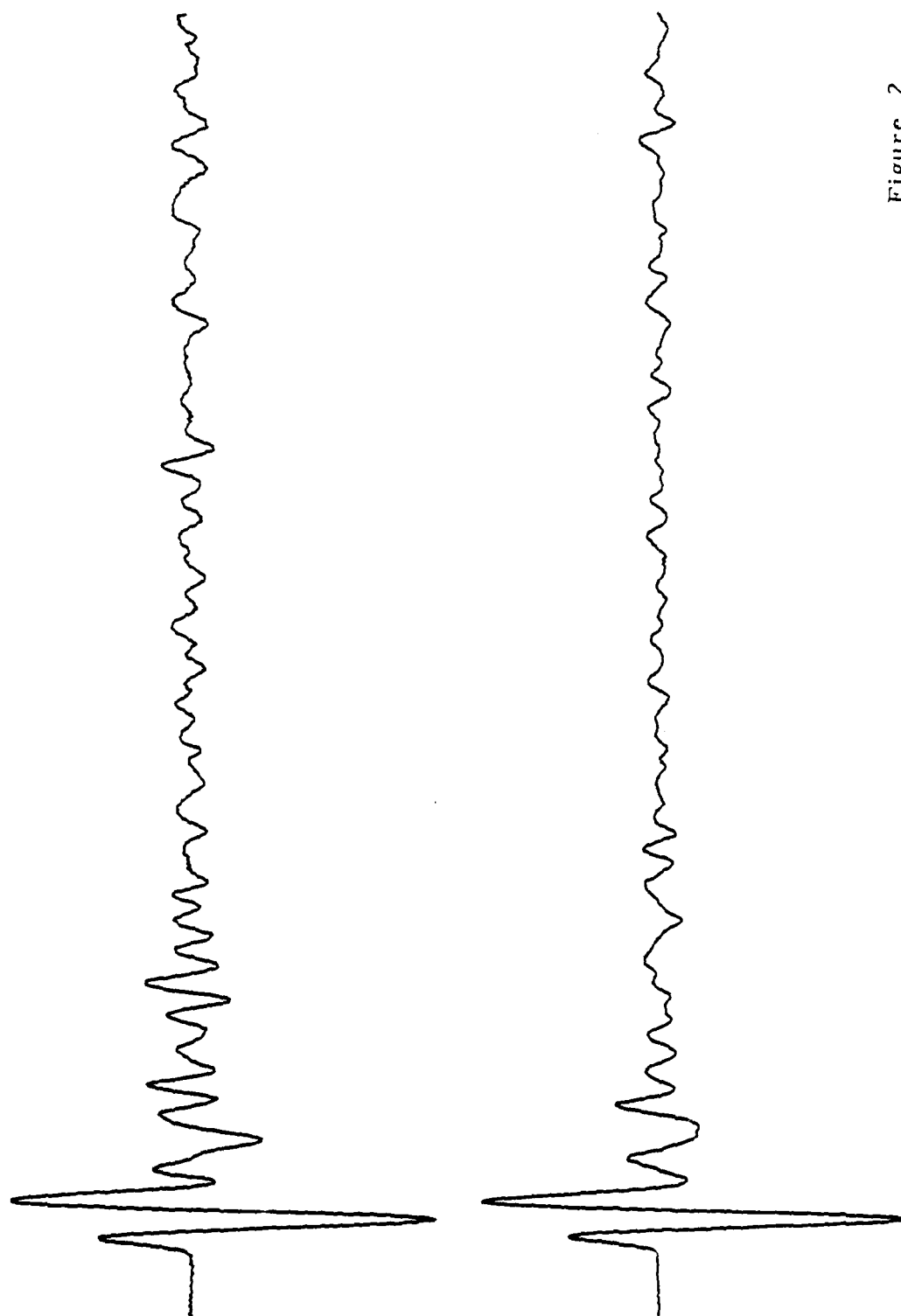


Figure 2

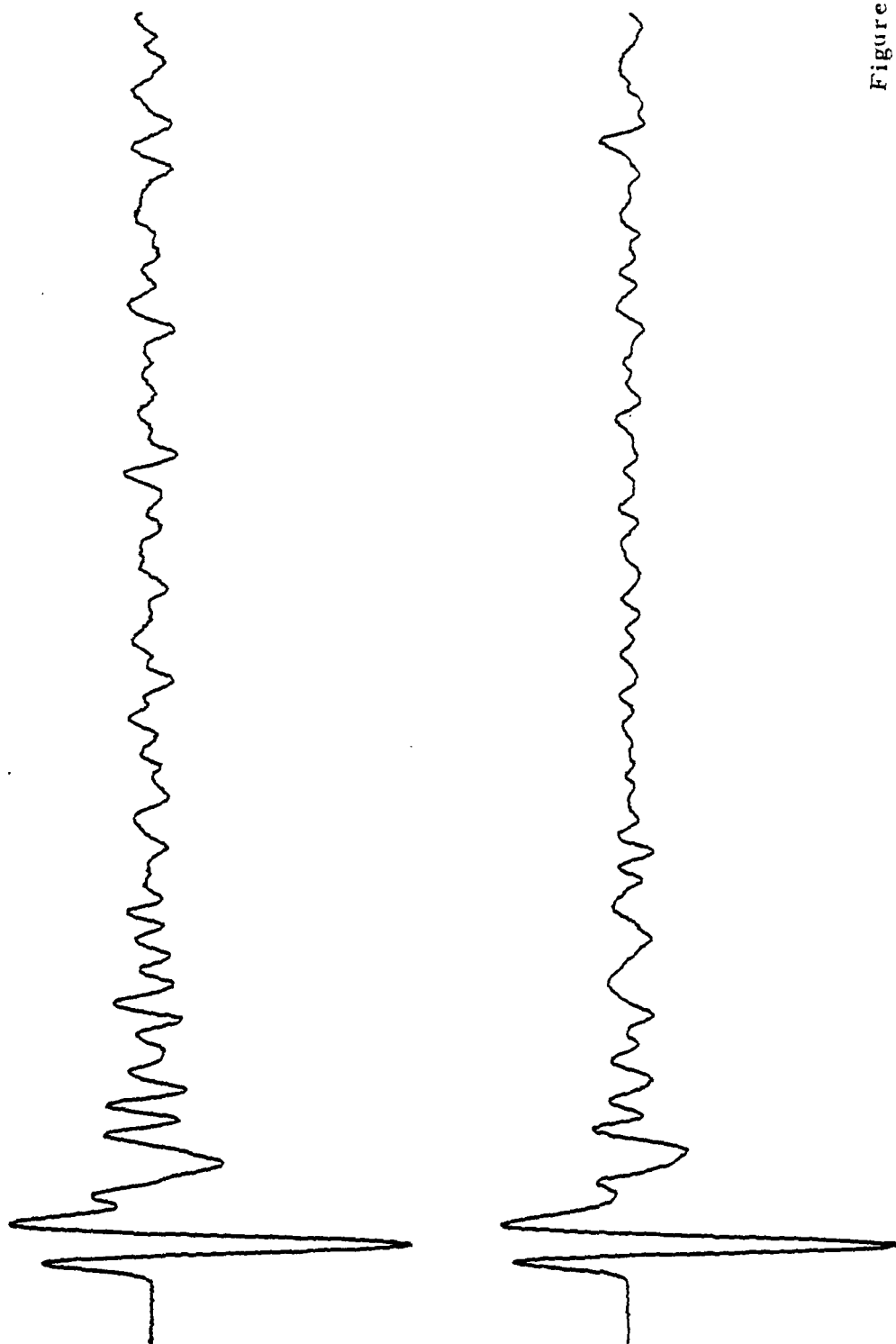


Figure 3

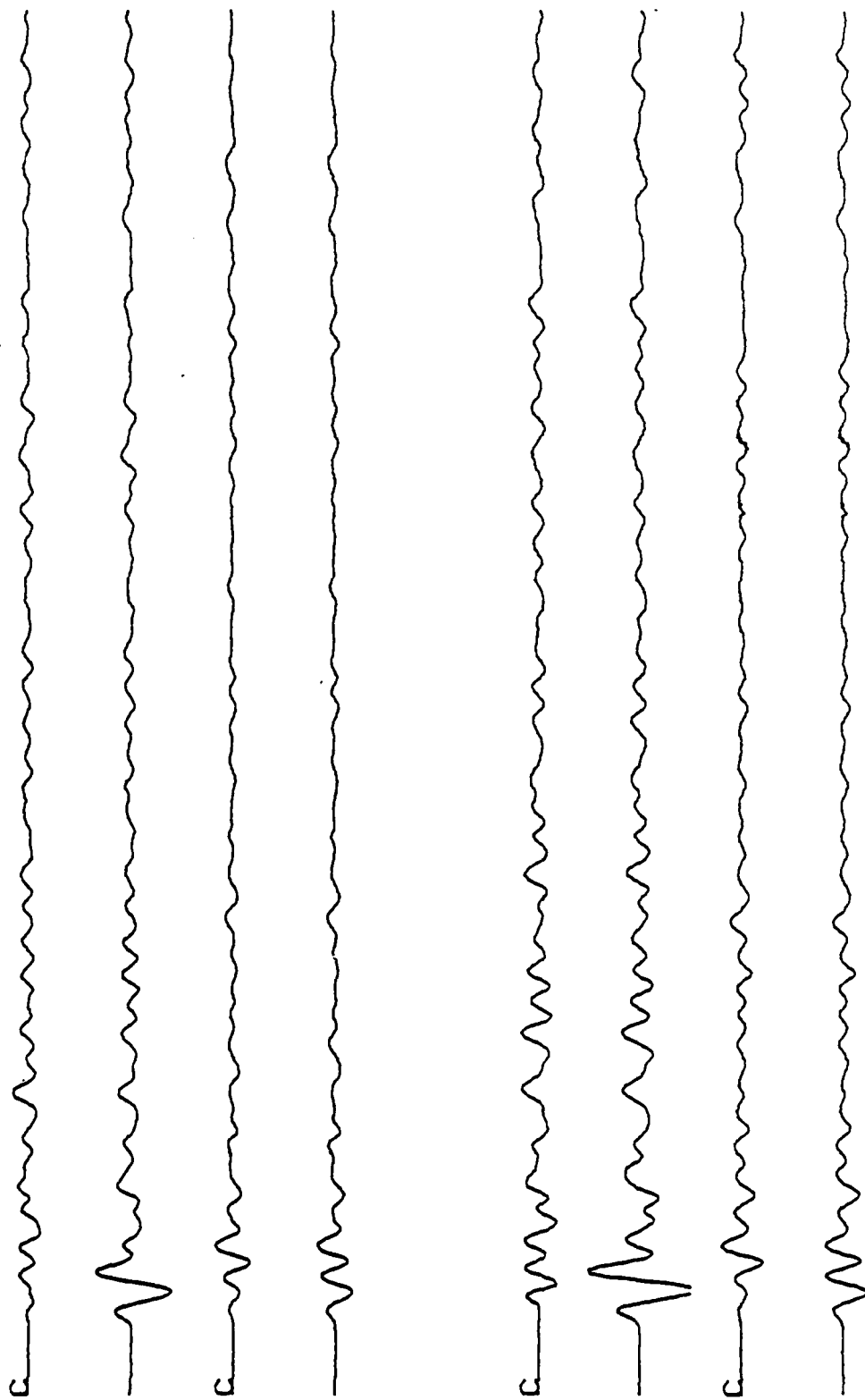


Figure 4 : Signals and Coda : for both events, original signals in two stations and the computed cauda : signal minus common signal (designated by C at the beginning).

Discussion

What could be considered as an advantage of a least square procedure is, that it always gives a solution. It is nevertheless very important to study the quality of the results.

At the first look, the differences between the plain sum of all signals and the commun signal are very small. Only the first onset of the records are not the same. But this is rather important. A large part of the information on the source mechanism is contained in this early part of the signal (in that case, we know the same mechanism to be very short). And so, our process does really increase the ability to study this mechanism.

We also computed the station corrections k_i for various time window length (the results plotted above are for the largest of the windows used). As long as the first two seconds of each record are in the window the k_i are stable (within 5%). For both quakes, the differences are less than 10%.

Future Work

Obviously we are only at the beginning of our work. We have to study the results of our data processing system :

- On a seismic swarm, to obtain better source function and to compare the source from one event to the others.
- On the array : to obtain station corrections versus azimuth, to apply the corrections obtained on a large event, on the records of smaller ones...
- On the study of the coda : others preliminary results let us think that within the coda, it is possible to define secondary arrivals. We wish to localize their origin.